

Earthquake Sequence 1973–1996 in Bárðarbunga Volcano: Seismic Activity Leading up to Eruptions in the NW-Vatnajökull Area

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Abstract — A day and a half after the earthquake ($m_b=5.3$, $M_S=5.6$, $M_W=5.6$) in the Bárðarbunga central volcano on Sept. 29th 1996, a volcanic eruption broke out under the Vatnajökull glacier. The eruption was located approximately 20 km SSE of the earthquake epicenter, midway between the Bárðarbunga and Grímsvötn central volcanoes. Course of events suggests a connection between earthquake and eruption and therefore a connection with a sequence of earthquakes of the same characteristics in Bárðarbunga during the years 1973–1996. The earthquakes in question are of an unusually low frequency character (corner frequency), explained by exceptionally low dynamic stress drop (< 10 bars) at shallow depth (≤ 5.0 km). The sequence which lasted for 22 years is characterised by \sim annual main events of magnitudes in the range of 4.5–5.7 (m_b). It intensified in the 1990s, with some of the largest earthquakes of the whole episode occurring at that time. Moment tensor solutions of teleseismic signals and locally recorded waveforms reveal that the main events are thrust faulting earthquakes with a significant non-double couple component. Arguments are presented that the faulting occurred on a steeply inward dipping caldera fault, with reactivated motion on a weak fault. As a consequence of this hypothesis magma inflation in Bárðarbunga is the most probable cause of the 1973–1996 events. However, the loading force (the magma) may or may not have resided at a similar shallow depth as the earthquakes. Cast in the frame of the inflation model, the Bárðarbunga 1973–1996 sequence implies a resurgent caldera of at least $0.2\text{--}0.7\text{ km}^3$ for approximately a quarter of a century, exceeding its magma storage capacity in 1996. However, these calculations are model dependent. Bárðarbunga and neighbouring area were relatively calm during the period mid-1997 to 2004. There was a renewed activity of small earthquakes during the years 2005–2009. From the beginning of continuous seismic recording in Iceland in 1925, all eruptions in Vatnajökull on record have been accompanied with earthquake(s) of magnitude ≥ 4.0 , within two months of the initial eruption.

INTRODUCTION

In 1973 an earthquake sequence started in Bárðarbunga, a central volcano under the North-Western part of Vatnajökull glacier (Figures 1 and 2). The sequence comprises a series of 20 main events of m_b magnitudes in the range of 4.5–5.7; shocks occurring once a year on average (Table 1). The last main event of this sequence occurred at 10:48:17.09 (GMT) on Sept. 29th 1996. A day and a half later an eruption broke out under Vatnajökull, ~ 20 km SSE of the main

event, midway between the Bárðarbunga and Grímsvötn central volcanoes. The eruption was named the Gjalp-eruption (Einarsson *et al.*, 1997; Guðmundsson *et al.*, 1997). The Bárðarbunga and Grímsvötn central volcanoes and their associated fissures are among the most active volcanic systems in Iceland. They have a history of causing volcanic disasters in the form of widespread poisonous gases, as well as producing the largest lava flow in historic time on earth, the 1783–1784 Laki eruption (the Skaftár fires), and large

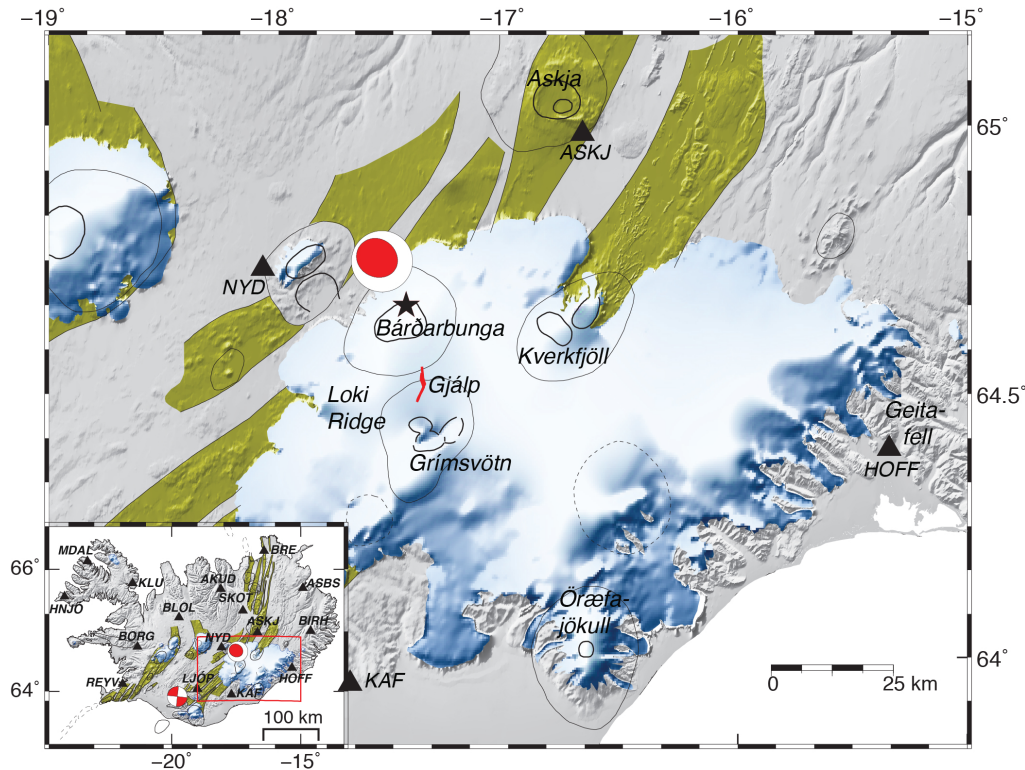


Figure 1. The Vatnajökull region. Volcanic centres are indicated by thin lines, and volcanic rift zones are in olive green shades (Einarsson and Sæmundsson, 1987). Inner rings of volcanic centres denote caldera rims. The star denotes the epicenter of the 5.6 (M_W) earthquake of Bárðarbunga in 1996. The moment tensor solution ('ball shape') indicates thrust faulting with a significant non-double couple component (Harvard University, USA). The bent red line denotes the Gjalp subglacial eruption fissure of 1996. Triangles show the location of the ICEMELT broadband seismic stations. Index map shows the volcanic rift zones, the ICEMELT network (Bjarnason *et al.*, 1996a) and focal mechanism of the Vatnafjöll earthquake of 1987. – Vatnajökulssvæðið. Virk eldstöðvakerfi og öskjubarmar eru táknud með þunnum línun, gliðnunar- og gosbelti eru ólíflugar (Einarsson og Sæmundsson, 1987). Stjarna sýnir staðsetningu skjálfta í Bárðarbungu árið 1996 af stærðinni 5,6 (M_W). Rauði og hvíti 'boltinn' táknar vægisþínu skjálftans (Harvard í BNA). Vægisþínan sýnir samgengishreyfingu á óvenjulegu misgengi eða margbrotna hreyfingu. Bogna rauða línan táknar staðsetningu sprungugossins í Gjalp árið 1996. Þríhyrningar sýna staðsetningu breiðbandsskjálftamæla í ICEMELT netinu (Bjarnason o.fl., 1996a). Á innsetningarkortinu kemur fram staðsetning ICEMELT netsins, lega gosbelta og brotlausn skjálftans í Vatnafjöllum árið 1987.

jökulhlaups from sudden volcanic melting of glacier ice. More frequently, however, did these volcanoes release energy in small eruptions without causing much harm to the community (Björnsson and Einarsson, 1990), even though the jökulhlaup which followed the 1996 eruption caused considerable damage to the road system in South Iceland (Haraldsson, 1997).

The proximity in time and space between the 1996 main event and the Gjalp eruption suggests a con-

nection between the two. The main objective of the present communication is to investigate the nature of the events described and the possible connection. The aim in particular is to analyse the source properties of the medium size earthquakes in the Bárðarbunga 1973–1996 earthquake sequence and other patterns in the seismicity that may relate to volcanism in the Northwest Vatnajökull during the 20th century.

Earthquake Sequence 1973–1996 in Bárðarbunga volcano

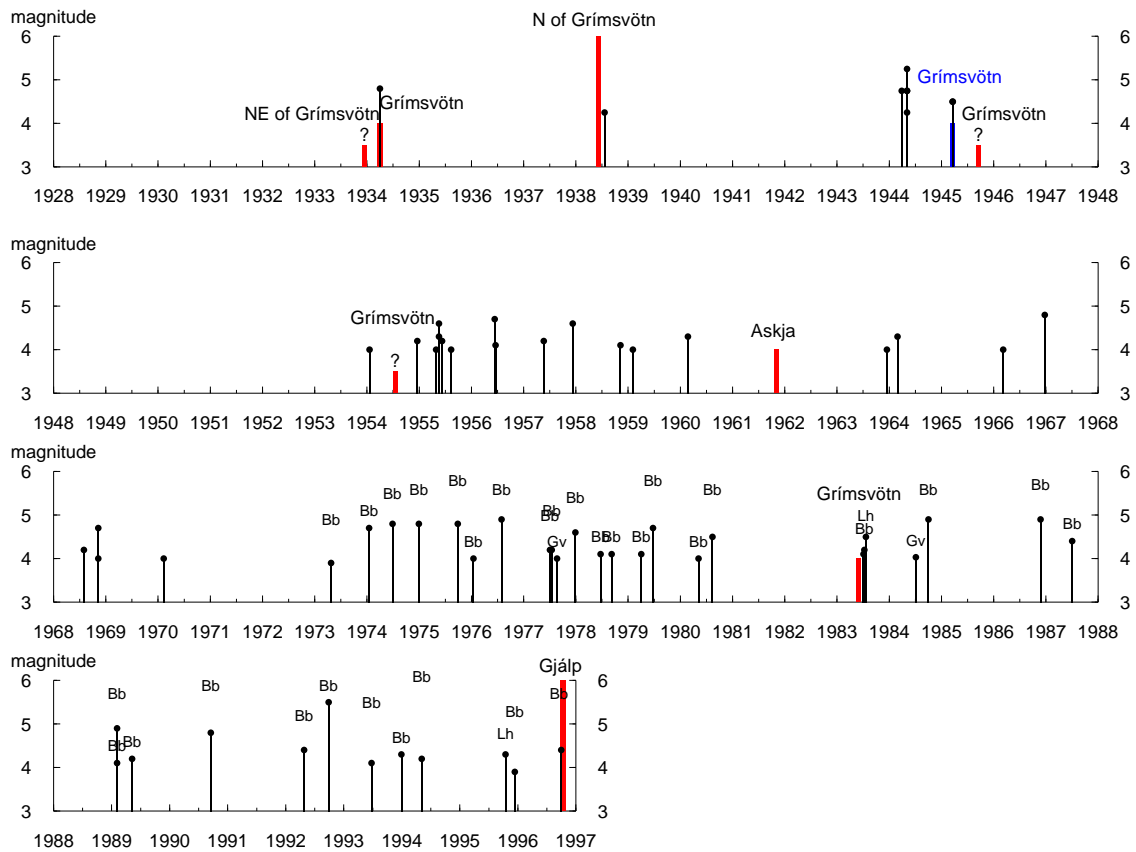


Figure 2. Magnitude time graph of earthquakes ($M_L \geq 4.0$) in the Vatnajökull region ($\sim 64.0\text{--}65.0^\circ\text{N}$, $15.0\text{--}18.5^\circ\text{W}$) during the period 1928 to 1996. Magnitudes are mostly local magnitudes. The earthquake time history is not complete for earthquakes of magnitude ≤ 4.4 for this region, prior to 1954, and locations are generally uncertain before 1974. Locations of the earthquakes are indicated relative to the nearest volcanic system, Bárðarbunga (Bb), Grímsvötn (Gv), and Loki Ridge (Lh). Red bars indicate eruptions in Askja, Grímsvötn, location north of Grímsvötn and Gjálp. Disputed eruptions (1933, 1945 and 1954) are indicated with question mark and the blue bar denotes proposed eruption in Grímsvötn on March 18-19th, 1945. Short bars represent small ($\leq 0.1 \text{ km}^3$) eruptions, or eruptions with unknown volume. Long bars represent eruptions with volume $> 0.1 \text{ km}^3$. Volume estimates from Einarsson (1962) and Guðmundsson (2005). Earthquake data from Tryggvason (1978a, 1978b, 1979), Ottósson (1989), Björnsson and Einarsson (1990), and ISC (2012). – *Tímaruna jarðskjálfta ($M_L \geq 4.0$) innan Vatnajökulssvæðisins ($\sim 64.0\text{--}65.0^\circ\text{N}$, $15.0\text{--}18.5^\circ\text{V}$) frá 1928 til 1996. Jarðskjálftasagan er ekki tæmandi fyrir 1954, að því er varðar jarðskjálfta ≤ 4.4 að stærð, og jarðskjálftastæðsetning er óviss fyrir 1973. Flestar stærðir jarðskjálftanna eru staðbundnar (M_L). Stæðsetning skjálfta er táknud með heiti eldstöðvakerfis, sem næst er skjálftanum, Bárðarbunga (Bb), Grímsvötn (Gv), og Lokahryggur (Lh). Rauðar línur tákna eldsumbrot í Öskju, Grímsvötnum, Gjálp og undir jökli norðan Grímsvatna. Óstaðfest eldsumbrot (1933, 1945 og 1954) eru táknud með stuttum rauðum línum og spurningarmerki, og blá lína er ný tilgáta um eldgos í Grímsvötnum dagana 18.-19. mars, 1945. Stuttar rauðar línur án spurningarmerkis tákna eldgos með litlu gosfni ($\leq 0.1 \text{ km}^3$) eða óþekktu gosmagni. Langar rauðar línur tákna eldgos með gosfni $> 0.1 \text{ km}^3$. Upplýsingar um gosfniðmat eru frá Einarssyni (1962) og Guðmundssyni (2005). Jarðskjálftagögn eru frá E. Tryggvasyni (1978a; 1978b; 1979), K. Ottóssyni (1989); H. Björnssyni og P. Einarssyni (1990); og ISC (2012).*

SEISMIC ACTIVITY AND VOLCANIC ERUPTIONS IN THE VATNAJÖKULL REGION

The early seismic net, sensitivity analysis

In Iceland continuous recording of seismic activity began (Reykjavík) in 1925 (Tryggvason, 1978a). In 1951–1954 a new and a more sensitive seismometer (Sprengnether) was placed in Reykjavík and the old one (Mainka) was transferred to Akureyri in North Iceland (Tryggvason, 1973). In the early days based on the Reykjavík recordings, epicenter locations were rather inaccurate for the distant Vatnajökull area. With the instrument upgrading of the 1950s, detection of earthquakes of magnitude 3.0 and greater improved for the Vatnajökull region, but location determinations were still inaccurate. Locations of small earthquakes in Vatnajökull were not well constrained until mid-1970s after the instalment of seismographs in North-east and later in East and Southeast Iceland (Einarsson, 1991). From the beginning of recording, the local magnitude scale (M_L) of Icelandic earthquakes is thought to have remained rather uniform (Tryggvason, 1973), except possibly after the installment of a digital national network in the 1990s (see Table 1).

Since 1954 the earthquake bulletins show increase in earthquakes of magnitude ≥ 3.0 in the Vatnajökull region (Tryggvason, 1979). Previously, authors came to the conclusion that the detection threshold for the Vatnajökull region in the years 1925–1953 was as good as $M_L \geq 3.5$ –4.0 (Tryggvason, 1973; Brandsdóttir, 1984). In good weather conditions the detection threshold will certainly improve significantly, but in bad conditions it deteriorates. Therefore, to interpret the natural changes in seismic activity in the Vatnajökull region in the 20th century, it is necessary to evaluate the magnitude of completeness (M_c) of the seismic bulletins preserved. M_c is a fixed value of an earthquake magnitude which subdivides a seismic bulletin into two parts. 95% of earthquakes of magnitude $>M_c$ will have a record in the bulletin. Many earthquakes that occurred in the region with magnitude $\leq M_c$ will, however, not have a record. Assessment of M_c for early recorded earthquakes in Iceland seems not have been carried out before.

The magnitude frequency relation of Gutenberg

and Richter can be used to make a preliminary estimate of the M_c value for the Vatnajökull region. Tryggvason (1973) reported relatively high b-values, 1.2–1.3, in the Vatnajökull and Dyngjufjöll area (i.e. where Askja is located in Figure 1) compared to other parts of Iceland. In the early years, 9 earthquakes can be interpreted to be within the Vatnajökull region in the range $M_L = 4^{1/4}$ – $5^{1/4}$. Based on these observations the M_c is estimated to be in the range 4.4–4.5 (M_L), assuming the Tryggvason (1973) b-values, and using the Gutenberg and Richter magnitude frequency relationship. If the global average b-value of unity would be assumed for Vatnajökull region, a lower M_c of $\sim 4^{1/4}$ would be expected.

In the early decades of seismic recording, earthquakes in Iceland were often detected at seismic stations outside the country, i.e. in Greenland or in continental Europe, even sometimes as low as 4.0 in magnitude (Tryggvason, 1978a, 1978b). Hence, with the local $M_c \sim 4.4$, and relatively good detection outside Iceland, it is unlikely that earthquakes of magnitude ~ 5.0 , like those of the 1973–1996 Bárðarbunga earthquake sequence, would have escaped detection in the early years of recordings.

Tryggvason (1973) concluded that the observed increased rate of earthquakes since 1955 in the Vatnajökull area to be real, and Björnsson and Einarsson (1990) pointed out a possible correlation between this increase in seismicity and increased geothermal activity in the Loki Ridge glacier cauldrons at the same time (Figure 1). Assuming $M_c \sim 4.4$ (M_L) in the early period, there is uncertainty if there is in fact a real increase in the rate of earthquakes with $M_L > 4.4$ in Vatnajökull region since 1955. However, the small number of earthquakes in the Vatnajökull area reported in the early period, limits what can be concluded.

From the mid-1970s the ability to locate epicenters became accurate enough in the Vatnajökull region, making it possible to map distinctly the seismically active areas in the region (Björnsson and Einarsson, 1990; Einarsson, 1991). The density of the permanent seismic network was not high enough to constrain sufficiently earthquake depth. However, the network did indicate that the majority of the seismic-

ity was shallow (<10 km) (Björnsson and Einarsson, 1990). Einarsson (1991) noted that earthquakes in the volcanic zone in the Vatnajökull region do not delineate major plate boundaries, but cluster on the central volcanoes, i.e. on Bárðarbunga, Grímsfjall, Hamarinn and Kverkfjöll, as well as the east-west trending volcanic Loki Ridge (Lokahryggur in Icelandic) (Figure 1). Of these, Kverkfjöll have been the least seismically active since the 1970s. The seismicity also shows that the stratovolcano Öraefajökull, which lies south of the volcanic zone, has been even less seismically active than Kverkfjöll. The seismic clustering around the Vatnajökull volcanoes has been explained partly by stress changes associated with a deflating magma chamber in the case of Bárðarbunga (Einarsson, 1991) and an inflating magma chamber in the case of Grímsvötn (Einarsson and Brandsdóttir, 1984). Although the seismicity does not delineate the major plate boundary faults, they may align on smaller faults. The seismicity around Bárðarbunga does e.g. delineate an arch shaped structure that approximates the caldera rim fault (Björnsson and Einarsson, 1990).

Eruptions and seismicity, short term correlation?

In the period 1934–1996 there have been 4 confirmed eruptions in Vatnajökull and one small eruption (0.1 km^3) in the Askja volcano, 25 km north of Vatnajökull (Einarsson, 1962; Jóhannesson, 1983; Björnsson and Einarsson, 1990) (Figure 2). Seismic tremor signals associated with jökulhlaup from sub-glacial geothermal areas in the Loki Ridge area were interpreted as small subglacial eruptions (Þorbjarnardóttir *et al.*, 1997; Einarsson *et al.*, 1997), but their certainty has not been confirmed by other methods. Three additional eruptions suggested in Grímsvötn in the years 1933, 1945 and 1954, respectively (Jóhannesson, 1983, 1984), have been disputed (Gudmundsson and Björnsson, 1991).

An intense earthquake swarm was recorded on the Reykjavík seismometers on March 18–19th 1945. Based on S–P time, the location of this event was traced to the Vatnajökull region (Tryggvason, 1978b). The largest of these earthquakes was $4\frac{1}{4} M_L$. It was followed by a single earthquake on March 20th ($M_L=4\frac{1}{4}$) in similar location. An experienced Vatna-

jökull traveller visited Grímsvötn in the summer of 1944 and again in July 1945. In July 1945 he observed at the southern rim of the subglacial Grímsvötn caldera a new opening with turbulent boiling water in the ice cover that floats on the subglacial lake (Þórarinnsson, 1974). No signs of such activity were seen the year before. Two months after his visit in July 1945, a jökulhlaup burst out from Grímsvötn. Björnsson and Guðmundsson (1993) have estimated the thermal output of the Grímsvötn caldera between 1922–1991. During this period the largest heatflux in Grímsvötn is associated with the 1934 Grímsvötn eruption. The second largest heatflux pulse in Grímsvötn, according to this report, occurred in the years 1945–1948. From the observations described, it is proposed, that the earthquake swarm of March 18–19th in 1945 signifies the onset of an eruption in Grímsvötn, which was never directly observed.

All confirmed eruptions in Vatnajökull in 1934–1996 were accompanied by earthquakes of magnitude 4 or more within 2 months of the beginning of eruptions (Figure 2). The last 3 eruptions in Grímsvötn (1998, 2004 and 2011, respectively) were all accompanied by events of magnitude 4.0 or larger (M_L or m_b) on the first day of eruption. During the eruption of Grímsvötn in 1983, earthquakes of this magnitude did, however, not happen (ISC, 2012). The reciprocal relationship does not hold, i.e. not all earthquakes of magnitude 4 and greater are associated with eruptions within the Vatnajökull region. As many of the eruptions in the Vatnajökull area are accompanied by earthquakes smaller than M_c (see definition above), the disputed eruptions of Grímsvötn in 1933, 1945 and 1954, cannot be rejected on the ground of the seismic bulletins available.

Subsurface pressure connection between volcanoes in Iceland and elsewhere has been postulated (Einarsson, 1991; Gonnermann *et al.*, 2012). Therefore, it is tempting to ask if volcanic events occurring close to Vatnajökull may have stimulating seismic effects on the Vatnajökull region. The neighbouring Askja eruption in 1961 does not seem to have induced seismicity at the level > 4.0 in Vatnajökull. On the contrary Vatnajökull remained seismically quiet at the time (Figure 2).

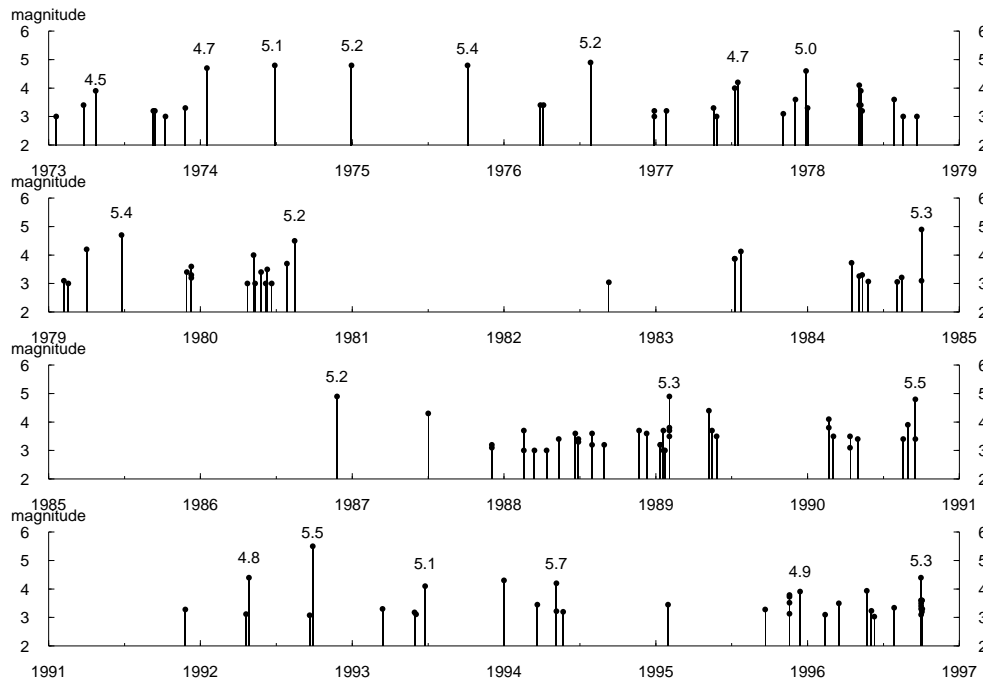


Figure 3. Magnitude time graph of earthquakes ($M_L \geq 3.0$) in the Bárðarbunga region ($\sim 64.5\text{--}64.8^\circ \text{ N}$, $17.0\text{--}17.8^\circ \text{ W}$) during 1973–1996. Magnitudes are mostly local (M_L), but listed (PDE-USGS) m_b magnitudes of main events are also labeled (see Table 1). Note, that a period of foreshock activity is not uncommon, but aftershocks are rare, suggesting an efficient stress release of the main events, possibly because of low stress environment. Earthquake data are from Björnsson and Einarsson (1990), USGS (1999) and ISC (2012). – *Tímaruna jarðskjálfta ($M_L \geq 3.0$) á Bárðarbungusvæðinu ($\sim 64.5\text{--}64.8^\circ \text{ N}$, $17.0\text{--}17.8^\circ \text{ V}$) frá árunum 1973–1996. Lengd lína gefur til kynna staðbundnar stærðir jarðskjálftanna, en m_b stærðir meginskjálfta eru auk þess sýndar með tölustöfum (sjá Töflu 1). Ekki er óalgengt, að forskjálftar fylgi meginskjálftum, en eftirskjálftar eru óalgengir (undantekning er 1996 skjálftinn). Skortur á eftirskjálftum bendir til skilvirkrar spennulosunar meginskjálftans, enda spenna tiltölulega lág við upphaf hans. Jarðskjálftagögn frá H. Björnssyni og P. Einarssyni (1990), USGS (1999) og ISC (2012).*

The Bárðarbunga 1973–1996 earthquake sequence and other seismic activity in the Northwest Vatnajökull region

The sequence of earthquakes that started in Bárðarbunga in 1973 can be described as a series of main events with magnitudes in the range 4.5–5.7 m_b and associated seismicity, occurring \sim yearly (Figure 3 and Table 1). The majority (2/3) of the main events are in the magnitude range of 5.0–5.7 m_b , but some as small as 4.5 m_b are defined as main events on the basis of the associated seismicity. By definition, a main event (mainshock) is the largest earthquake in a sequence of earthquakes close in space and

time, the so-called pattern of fore-, main-, and aftershocks. Foreshocks have been recorded with many of the Bárðarbunga main events, but the sequence is unusual in that it lacks significant aftershocks (Einarsson *et al.*, 1997). An exception to this pattern was the 1996 main event ($M_W=5.6$). There active fore- and aftershock sequences happened. The interval between main events has varied. The seismic activity was significantly less in the 1980s (interval between main events ~ 2.5 years) compared with the 1970s and 1990s (interval ~ 1.0 year). Body-wave magnitudes (m_b) have been determined for all 20 main events of the sequence (Table 1). A comparison of the m_b magnitudes shows that although they are on average sim-

Table 1. Listings of Bárðarbunga main earthquakes 1973–1996. – *Meginjarðskjálftar í Bárðarbunga*.

| <i>ár</i> | Date | <i>dags.</i> | time UT <i>tími</i> | latitude <i>n.l. breidd</i> | longitude <i>v.l. lengd</i> | M_L | m_b | obs ^a | M_S | obs ^a | M_W |
|-----------|------|--------------|------------------------|--------------------------------|--------------------------------|-------|-------|------------------|-------|------------------|-------|
| 1973 | Apr | 23 | 02:57:16 | 64.6 | -17.3 | 3.9 | 4.5 | | | | |
| 1974 | Jan | 15 | 19:47:30.8 | 64.63 | -17.34 | 4.7 | 4.7 | | | | |
| 1974 | June | 25 | 22:23:41.1 | 64.71 | -17.53 | 4.8 | 5.1 | 27 | 5.2 | 2 | |
| 1974 | Dec | 29 | 03:50:00.1 | 64.71 | -17.42 | 4.8 | 5.2 | 31 | | | |
| 1975 | Oct | 03 | 18:34:03.3 | 64.439 | -17.285 | 4.8 | 5.4 | | | | |
| 1976 | July | 27 | 04:00:51.0 | 64.654 | -17.406 | 4.9 | 5.2 | | 4.9 | | |
| 1977 | July | 14 | 07:15:31.7 | 64.622 | -17.403 | 4.2 | 4.7 | | | | |
| 1977 | Dec | 28 | 20:32:38.6 | 64.671 | -17.372 | 4.6 | 5.0 | | 5.2 | | 5.4 |
| 1979 | June | 22 | 23:17:57.8 | 64.615 | -17.396 | 4.7 | 5.4 | | 4.8 | | 5.2 |
| 1980 | Aug | 12 | 12:11:42.9 | 64.646 | -17.352 | 4.5 | 5.2 | | 5.3 | | 5.5 |
| 1984 | Sept | 30 | 23:31:53.5 | 64.607 | -17.403 | 4.9 | 5.3 | 45 | 4.6 | 5 | |
| 1986 | Nov | 23 | 02:48:59.7 | 64.661 | -17.333 | 4.9 | 5.2 | 49 | | | |
| 1989 | Feb | 03 | 15:18:23.1 | 64.595 | -17.406 | 4.9 | 5.3 | 58 | 4.9 | 3 | 5.1 |
| 1990 | Sept | 15 | 23:07:41.1 | 64.65 | -17.39 | 4.8 | 5.5 | 58 | 5.2 | 13 | 5.6 |
| 1992 | Apr | 25 | 06:48:43.9 | 64.725 | -17.482 | 4.4 | 4.8 | 58 | 4.4 | 6 | |
| 1992 | Sept | 26 | 05:45:50.45 | 64.646 | -17.537 | | 5.5 | 99 | 5.4 | 45 | 5.7 |
| 1993 | June | 22 | 12:33:43.67 | 64.680 | -17.180 | 4.1 | 5.1 | 76 | 4.8 | 26 | 5.3 |
| 1994 | May | 05 | 05:14:48.55 | 64.634 | -17.430 | 4.2 | 5.7 | 91 | 5.2 | 45 | 5.4 |
| 1995 | Dec | 11 | 05:22:46.19 | 64.671 | -17.504 | 3.9 | 4.9 | 72 | 4.5 | 4 | |
| 1996 | Sept | 29 | 10:48:17.09 | 64.666 | -17.444 | 4.4 | 5.3 | 78 | 5.4 | 37 | 5.6 |

Locations and M_L are from IMO (2015), except for the Sept. 26th, 1992 (ISC, 2012) and the Sept. 29th, 1996 event (Einarsson *et al.*, 1997). M_S , m_b and ^a number of observations from USGS (1999). M_W are determined by Harvard University, Cambridge, USA.

ilar during the three decades, they are 4–5% larger in the 1990s. The moment magnitude (M_W) that correlates best with the energy release of earthquakes has been determined for 9 of the earthquakes in the Bárðarbunga sequence. The three highest moment magnitude events all occurred in the 1990s, and the Sept. 1996 event was one of them. There is thus an indication of intensified seismic activity of the sequence in the 1990s.

Prior to the Sept. 1996 earthquake, there had been a highly active seismic period for about one year in Bárðarbunga and the surrounding area (Figures 2 and 3). An earthquake swarm occurred within the Hamarinn volcano in February 1996 (Einarsson *et al.*, 1997). Seismic tremor accompanied two jökulhlaups from the Skaftár cauldrons in July 1995 and August 1996. These tremors had characteristics of eruption tremor and have been interpreted as such, implying small sub-glacial eruptions lasting 1/2–2 days on the volcanic Loki Ridge (Þorbjarnardóttir *et al.*, 1997).

The combination of these events suggests an increased pressure over large region of Northwest Vatnajökull prior to the Gjálp eruption in Sept. – October of 1996. The subsequent eruption in Gjálp may have caused pressure drop in Bárðarbunga and neighbouring regions, although initially there was increased

seismic activity in the area of Loki Ridge and Hamarinn until middle of year 1997 (Figure 4 in Jakobsdóttir, 2008). There was a relative quiescence in seismicity of earthquakes of magnitude ≥ 3.0 from June 1997 to 2005 in the Northwest part of Vatnajökull, including the most active volcanoes Bárðarbunga, Loki Ridge and Hamarinn (ISC, 2012; IMO, 2015; Figure 4 in Jakobsdóttir, 2008). The Grímsvötn eruption of 1998 and 2004 may also have influenced the quiescence observed.

Earthquakes with hypocenters in the depth range 20–30 km are uncommon in Iceland, and can be described as deep earthquakes relative to the common seismicity. Interestingly in the years 2005 to 2009 large portion of the deep seismicity in Iceland occurred in the Bárðarbunga region (64.5–64.8°N, 17.0–17.8°W; IMO, 2015). The closest stations used for the location of the deeper seismicity are within 40–50 km and S phases are generally used in their locations (Martin Hensch, *pers. comm.*, Feb. 2015). Gomberg *et al.* (1990) showed that a good constrain on focal depth is generally obtained given correctly timed S phase recorded within ~ 1.4 focal depth distance from the epicenter of an earthquake. The IMO network is therefore on the border of fulfilling this requirement for the deeper Bárðarbunga events.

In the routine location of earthquakes in the Vatnajökull region the IMO uses velocity models with ratio $V_P/V_S=1.78$. Combining the velocity models of Darbyshire *et al.* (1997) and Bjarnason and Schmeling (2009) for Central Iceland, there is indication that a ratio $V_P/V_S \sim 1.85$ may be more appropriate for the closer stations in the Central Iceland region. Such higher V_P/V_S ratio would, however, tend to make the located depth of the earthquakes shallower. If the depths of these deeper Bárðarbunga earthquakes can be constrained, even at shallower depth than currently located (i.e. >15 km), it will be postulated that deeper earthquakes under Northwest Vatnajökull are caused by fracturing of rocks around the crust-mantle boundary and lower crust by ascending magma from the mantle, as occurred during the Westman Islands eruption in 1973. There, however, a temporary dense seismic net was installed during the eruption and recorded well constrained earthquakes at 15–25 km depth under the eruption site, which are explained by magma induced strain release (Björnsson and Einarsson, 1981; Einarsson 1991). It is unknown if similar deep earthquakes may have occurred in the 1973–1996 earthquake sequence. However, the Science Institute of University of Iceland operated analogue seismic stations in Central Iceland during good part of the years 1973–1996. One of these stations was located within the 1.4 focal depth distance of potential deep earthquakes in the Bárðarbunga area, as well as one station in the ICEMELT network in the years 1995–1996. It is conceivable, that the question regarding deep earthquakes under Bárðarbunga can be answered by analysing these old data. Monitoring deep earthquakes under volcanic systems in Iceland could become an important tool for volcanic hazard prediction within intermediate time frame (years to decades).

Focal mechanisms have been constructed for a number of the Bárðarbunga earthquakes. The mechanisms indicate thrust faulting with a strike-slip component, with vertical or sub-vertical T-axis (Einarsson, 1991). Moment tensor solutions show also thrust faulting with a significant non-double-couple component (Ekström, 1994; Nettles and Ekström, 1998; Tkalčić *et al.*, 2009). Rifting and transform are the

predominant tectonic motions in Iceland (Sæmundsson, 1979), and thrust faulting as indicated by small and medium size earthquakes is not often observed in the surface tectonics of the country (Gudmundsson *et al.*, 2008).

In 1994–1996 the ICEMELT digital broadband seismic network (Bjarnason *et al.*, 1996a; 1996b) recorded the last three main events of the Bárðarbunga sequence (Figure 4). Waveforms were similar in all of them and characterised by emergent P waves and large amplitude surface waves. The 1996 event was clearly the largest of the three. They have in common low corner frequency compared to a number of earthquakes in Iceland of similar size that have been examined (Table 2). The 1996 event has the lowest P wave corner frequency (0.17 ± 0.03 Hz) of the three events with very low frequency P waves (≈ 0.2 Hz) arriving approximately 0–3 s after the first motion (Appendix, Figures A1 and A2). These low frequency P waves are likely to be produced at or near to the source, as they clearly arrive with the first motion on many of the ICEMELT stations (Figure A2).

At the time of the 1996 main event in Bárðarbunga, the closest seismic station in the IMO-network to the epicenter was at ~ 100 km distance (Jakobsdóttir, 2008). Hence the hypocenter depth of the 1996 earthquake and most previous Bárðarbunga main events are currently in general unconstrained in the seismic bulletins, but some improvements may be possible. All moment tensor inversions of the 1996 main event find the best fitting centroid depth at 3.5 km (Nettles and Ekström, 1998; Konstantinou *et al.*, 2003; Tkalčić *et al.*, 2009). In the following section it will be argued that Bárðarbunga main events are unusually shallow (<5 km) for earthquakes of intermediate size. Such a shallow depth for intermediate earthquakes is unusual in a global perspective, and can account for the unusual source properties of the sequence.

DISCUSSION

The Bárðarbunga sequence is reasonably well documented, but the forces behind it are not well understood. One can speculate that it is either of plate tectonic or of localised magmatic origin, although the

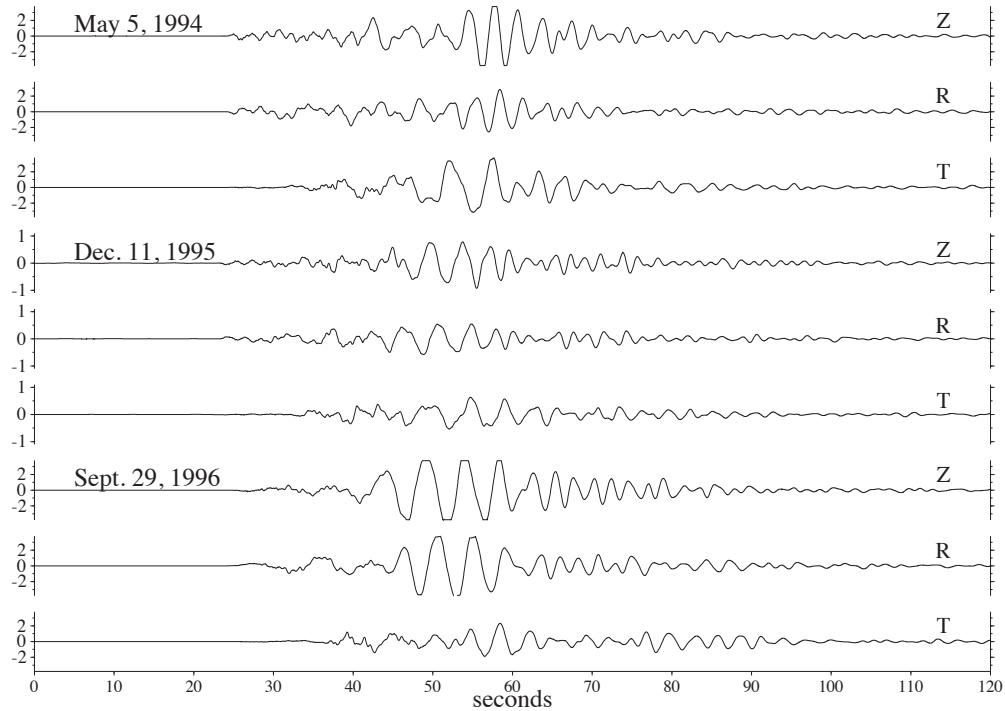


Figure 4. Waveforms of Bárðarbunga earthquakes 1994, 1995, and 1996, recorded by the ICEMELT broadband station KAF, in SE-Iceland (Figure 1). Two minutes long, 3-component, vertical, radial and transverse (Z, R, T) records are shown for each event. Note that the amplitude scale is not the same for all records. Some of the surface wave phases of the 1994 and 1996 events are clipped, but the 1996 event is obviously the largest of the three. – *Bylgjugögn þriggja megin skjálfta í Bárðarbungu 1994, 1995, og 1996, frá breiðbandsmæli ICEMELT netsins á Kálfafelli (KAF), Suðausturlandi (sjá 1. mynd). Tímagluggi er tveggja mínútna langur, og þrír þættir hraðahreyfingar jarðar eru sýndir í lóðrétt- (Z), geisla- (R) og þverstefnu (T). Athugið, að skali á bylgjuútslagi (y-ás) er ekki sá sami fyrir skjálftana. Sumar yfirborðsbylgjur skjálftanna frá 1994 og 1996 eru skornar (mælar hafa farið í botn), en 1996 skjálftinn er greinilega stærstur þessara skjálfta.*

Table 2. Examples of small and medium size earthquakes in Iceland. – *Dæmi um jarðskjálfta á Íslandi.*

| Date days. | time UT tími | latitude n.l. breidd | longitude v.l. lengd | m_b | M_S | M_W | f_c , ^a Hz | f_c , ^b Hz | obs ^c | region |
|---------------|-----------------|-------------------------|-------------------------|-------|-------|-------|----------------------------|----------------------------|------------------|---------------------------|
| 1987 May 25 | 11:31:54.7 | 63.909 | -19.779 | 5.8 | 5.8 | 5.9 | 0.50 | ±0.1 | 1 | Vatnafjöll ¹ |
| 1994 Feb. 08 | 03:27:52.1 | 66.451 | -19.249 | 5.3 | 5.3 | 5.5 | 0.43 | ±0.06 | 3 | Skagafjörður ² |
| 1994 May 05 | 05:14:48.6 | 64.634 | -17.430 | 5.7 | 5.2 | 5.4 | 0.22 | ±0.04 | 3 | Bárðarbunga ² |
| 1994 Aug. 20 | 16:40:25.9 | 64.035 | -21.241 | 4.3 | | | 3.2 | ±0.3 | 2 | Hengill ² |
| 1994 Nov. 18 | 23:54:03.7 | 64.508 | -17.686 | 4.2 | 3.9 | | 1.2 | ±0.4 | 3 | Loki Ridge ² |
| 1995 Oct. 13 | 15:14:16.0 | 64.484 | -17.749 | 4.3 | 3.9 | | 1.4 | ±0.5 | 4 | Loki Ridge ² |
| 1995 Nov. 18 | 01:56:23.9 | 64.647 | -17.419 | 4.3 | | | 0.9 | ±0.3 | 4 | Bárðarbunga ³ |
| 1995 Dec. 11 | 05:22:46.2 | 64.671 | -17.503 | 4.9 | 4.5 | | 0.47 | ±0.13 | 4 | Bárðarbunga ² |
| 1996 Sept. 29 | 10:48:17.09 | 64.666 | -17.444 | 5.3 | 5.4 | 5.6 | 0.17 | ±0.03 | 7 | Bárðarbunga ⁴ |

^aP wave corner frequency. ^bUncertainty of P wave corner frequency. ^cNumber of observations. Locations from: ¹Bjarnason and Einarsson (1991), ²SIL network, ³PDE of USGS, ⁴Einarsson *et al.* (1997). M_W are determined by Harvard University, Cambridge, USA, m_b and M_S by USGS.

two do not need to be decoupled in general. A plate-tectonic origin could be in the form of ridge-push driven tectonics, acting on the bend in the Central Iceland Rift at Bárðarbunga (i.e. the bend from SW-NE strike to a SSW-NNE strike) (Figure 1). There are several problems with this hypothesis. A ridge-push that generates thrust faulting would probably be larger outside the rift than within it, and evidence of regional compression has not been observed in Iceland (Khodayar and Einarsson, 2004). Rifting is also generally perceived as a passive process, a passive opening due to plate tectonics, that does not cause major thrust earthquakes as occur in Bárðarbunga.

The evidence of a magmatic origin of the Bárðarbunga medium size earthquakes comes from local seismic recordings. They tend to show relatively emergent P and S waves with low corner frequency and large surface waves that characterise volcanic earthquakes (e.g. Lahr *et al.*, 1994) (Table 2, Figure 4 and Appendix Figures A1 and A2). The tight clustering of the seismicity within the Bárðarbunga volcano also suggests magmatic origin (Einarsson, 1991).

Kinematics of the Bárðarbunga 1973–1996 earthquake sequence

It is argued in the present communication that during the largest earthquakes of the sequence ($M_W=5.4-5.7$), faults or fault patches of the size of $\sim 1/3$ of Bárðarbunga caldera circumference are being moved. The total circumference of the ice covered Bárðarbunga caldera is estimated to be ~ 30 km (Björnsson, 1988). These fault lengths can be inferred from the size of the earthquakes, assuming an average fault-slip-to-fault-length ratio to be 10^{-4} to 10^{-5} (Scholz *et al.*, 1986). The length of the 1996 fault can also be inferred from the distribution of aftershocks as located by Stefánsson *et al.* (1996). The whole aftershock sequence of the 1996 event is complicated, and it is likely to be generated by a number of processes, e.g. afterslip on the main fault, slip on neighbouring faults, and possibly lateral magma migration away from Bárðarbunga. However, in the first 12 hours after the main event of Sept. 29th 1996, many of the aftershocks were located in a relatively narrow arch shaped region that approximately follows the western half of the caldera rim, indicating that a ~ 12 km long fault segment rup-

tured on Sept. 29th 1996. The locations of the IMO (Stefánsson *et al.*, 1996) are not accurate enough to determine if the caldera fault moved, or a concentric fault to the west of it.

The Earth does often select pre-existing planes of weakness for stress release, even when they are definitely outside the plane of maximum shear stress. Cone sheets (inclined sheets) are common within eroded central volcanoes in Iceland (e.g. Annells, 1968; Sigurdsson, 1970; Fridleifsson, 1973; Jóhannesson, 1975; Torfason, 1979; Fridleifsson, 1983; Siler and Karson, 2009; Burchardt *et al.*, 2011, Gudmundsson *et al.*, 2014), and they may form arch shaped planes of weakness. However, the size of the Bárðarbunga earthquakes makes the caldera ring fault perhaps a more likely candidate. The third possibility is shear fracturing during the formation of a cone sheet. Anderson (1936) proposed that cone sheet formation is pure tension fracturing, but Phillips (1974) concluded that cone sheets occupy shear fractures. Gudmundsson (2002) studied thousands of cone sheets in Iceland. He concluded that they are primarily tension fractures. However, Torfason (1979) has documented in South-East Iceland several instances of shear movement across cone sheets usually with reverse sense of motion. Reactivated faulting on a cone sheet, or cone sheet formation, as a possible source of the Bárðarbunga earthquakes should therefore not be disregarded (see an excellent discussion by Shuler *et al.*, 2013). However, if cone sheets are pure tension fractures, then the moment tensor solutions of Bárðarbunga events (Nettles and Ekström, 1998; Konstantinou *et al.*, 2003; Tkalčić *et al.*, 2009) would probably rule out cone sheets as their main source.

A fault slip of an area 10 km by 3 km, with the shape of a steeply dipping cylindrical wall with centroid depths of 1.5 km and 5.0 km depth, respectively, allows us to calculate 23 to 66 cm and 13 to 36 cm slip for earthquakes of magnitude 5.4–5.7 (M_W), respectively. The assumed rigidity structure of Bárðarbunga for these calculations is derived from the V_{SV} velocity model of Bjarnason and Schmeling (2009) for Central Iceland (10 GPa and 32.5 GPa at 1.5 km and 5.0 km depths, respectively). Higher slip values would be obtained for cone shaped fault segments

with the same area, located at similar depths, because their centroid depth would tend to be shallower (i.e. higher proportion of the fault lies at a shallower depth with lower average rigidity). Assuming that displacements of the Bárðarbunga main earthquakes are accumulative, and by connecting their m_b and M_W magnitudes with a linear relationship, the accumulated moment can be calculated and related to the dimensions of the Bárðarbunga volcano. If for such an exercise, the total slip is distributed along the entire caldera rim fault (i.e. 30 km long and 1 km wide fault with centroid depth of 1.5 km), a ~ 9.5 meters total displacement is calculated with corresponding volume change of $\sim 0.7 \text{ km}^3$ of the caldera. If, however, the centroid depth is at the lower crustal boundary of 4–5 km depth (Darbyshire *et al.*, 1998; Bjarnason and Schmeling, 2009), the accumulated slip on the same fault geometry would be ~ 3.0 – 3.5 meters with a volume change of 0.2 – 0.25 km^3 . These numbers are likely to be minimum estimates, because earthquakes of lower size than 4.5 do not enter the calculation, and part of the deformation is most likely aseismic.

The above calculated volume changes can be compared with results from Árnadóttir *et al.* (2009) who carried out country wide GPS measurements in Iceland over the time period 1993 to 2004. These researchers observed a significant uplift (~ 8 – 18 mm/yr) of a broad area of Central and Southeast Iceland, which they modelled with glacial isostatic adjustments due to recent thinning of the largest glaciers in the country. In spite of relatively coarse GPS measurements around the Vatnajökull glacier, they do model a net $\sim 0.1 \text{ km}^3$ volume contraction under Bárðarbunga during the interval of observations, carried out in 1993 and 2004, respectively. Due to lack of temporal resolution in the GPS data, their study cannot resolve a possible variation in volume change before the 1996 Bárðarbunga main event and the Gjalp eruption, and a post eruption volume change, making comparison somewhat limited.

The moment tensors of the Bárðarbunga earthquakes that have been determined have a large non-double-couple component that could be consistent with earthquakes on circular faults or a collection of fault surfaces that form a circular assemblage (Ek-

ström, 1994; Nettles and Ekström, 1998; Konstantinou *et al.*, 2003; Tkalčić *et al.*, 2009). Nettles and Ekström (1998) and Tkalčić *et al.* (2009) interpret moment tensors to show that the main motion is subsidence on outward-dipping fault (with respect to the volcano), while Bjarnason and Þorbjarnardóttir (1996) interpreted the main motion to be an upward movement on inward dipping fault (Figure 5). Full moment tensor solution of Konstantinou *et al.* (2003) for the 1996 event resolved implosive isotropic component, with normal faulting, in contrast to the thrust faulting determined by Nettles and Ekström (1998), Tkalčić *et al.* (2009) and by Einarsson (1991) for previous Bárðarbunga events. However, Tkalčić *et al.* (2009) show with synthetic waveforms how data noise and slight error in velocity structure can lead to false isotropic component. Assuming that the main motion of the Bárðarbunga intermediate earthquakes is due to thrust faulting, then this motion can be interpreted as pure volcano deflation (Einarsson, 1991), or inflation (Bjarnason and Þorbjarnardóttir, 1996), or a combination of both (Nettles and Ekström, 1998; Tkalčić *et al.*, 2009). Without further information, i.e. on the dip of the faults, or detailed geodetic measurements of the volcano, which were not carried out at the time, a clear cause cannot be fully constrained. It is conceivable that the dip of the active faults can be determined, e.g., with relative locations of the 1996 aftershocks. However, the number of recorded aftershocks may not be high enough to allow for such an analysis. There is also uncertainty in identifying the true aftershocks. Other geological events may have occurred after the main event, e.g. formation of a ring dyke or cone sheets in the caldera area, inducing seismicity outside the main fault.

Structures of Icelandic calderas

Of paramount value is to gain information on the dip of the faults that have moved during the Bárðarbunga events, in order to understand the nature of their sources. Cone sheets generally dip towards the centre of volcanoes. There is a consensus that the drop of a piston type caldera (i.e. drop with relatively intact caldera floor) is accommodated on a steeply dipping near vertical ring fault. However, there is a lack of consensus on the general direction of the dip of the

main ring fault of calderas, or even on what is commonly observed in the field (e.g. Walker, 1984; Gudmundsson, 1998b; Roche *et al.*, 2000; Gudmundsson, 2007). Anderson (1936) predicted, theoretically, that calderas ring faults should dip steeply outward, with reverse sense of motion above an underpressured magma chamber.

The Anderson model has been favoured for a long time and has in general been confirmed by analogue experiments on caldera structures and development in the laboratory (e.g. Roche *et al.*, 2000; Burchardt and Walter, 2009). Acocella (2007) has reviewed the subject. Most of the work carried out in this field seems to comply with the Anderson-type ring fault in the first stages of caldera collapse (Acocella, 2007). In later collapse stages of caldera formation, the laboratory experiments show a second set of ring faults develop, with inward dip and normal sense of motion. The second set of ring faults was not a part of Anderson's prediction, perhaps because his analytical theory was describing the initial stress stages in caldera formation, and because his theory explained well observations of calderas in Scotland in his time. The second major ring fault is concentric with the initial major ring fault but lies further outside, which increases the diameter of the caldera. The analogue experiments indicate that inward dipping major fault has steeper dip than the initial major ring fault at shallow depth, but the two join at greater depth (Acocella, 2007). Burchardt and Walter (2009) have shown with analogue experiments that the drop of the caldera floor in later stages of a caldera development is increasingly taken up with normal faulting on the outer lying ring fault.

In studies on deeply eroded (1–2 km) calderas of extinct central volcanoes in Iceland, usually only one set of near vertical ring fault patches is reported (e.g. Sigurdsson, 1970; Fridleifsson, 1973; Jóhannesson, 1975; Torfason, 1979; Franzson, 1978). An exception is perhaps Geitafell, an extinct central volcano in Southeast Iceland (Figure 1), where patches of concentric sets of ring faults are observed inside and outside the main ring fault (Fridleifsson, 1983). None of the reported studies mentioned did directly measure the degree of the dip angle of the caldera fault. They usually infer inward dip with normal sense of

motion, based on a sharp change in the dip of the strata with steeply inward dipping layers, just inside the caldera fault. Recently, however, a measurement has been carried out on a 300 m long segment of caldera fault in Southwest Iceland that has an average 85° inward dip with normal sense of motion (Browning and Gudmundsson, 2015). Jóhannesson and Sæmundsson (2009) have mapped ~15 eroded calderas in Iceland. All of them have inward dipping ring fault patches, interpreted to be a part of the main ring fault of the calderas (H. Jóhannesson, *pers. comm.*, Jan. 2015, and several other geologists). So far the only exception found to this comes from Steffi Burchardt (*pers. comm.*, Feb. 2015). She observed curved outward dipping antithetic fault patches in the extinct Geitafell volcano, in the same outcrop as the main inward dipping ring fault. However, an antithetic fault may not be a good candidate for a major second set of a ring fault.

Fridleifsson (1983) mapped curved patches faults close (~1.0 km) to the Geitafell's main caldera fault. One of these, which can be traced a considerable distance, has inward dip with reverse sense of motion, and appears to join the caldera fault. Fridleifsson (1983) speculates that this fault first acted as a reverse fault, but later as caldera fault. In the present communication it is, however, proposed that the opposite may have happened. The reverse fault is reactivated caldera fault from the time of caldera resurgence.

For a normal fault, originally close to vertical, to be reactivated into a reverse fault, a major change in stress field is required. The question arises if there are more signs in the geological record of Iceland than already reported (Fridleifsson, 1983; Gudmundsson *et al.*, 2008), to support that normal faults have been reactivated. Sibson (1985) presents an expression for the optimal angle (θ^*) between a fault plane and the maximum principal stress (σ_1) for reactivation, that depends on the coefficient of friction (f). For faults of normal strength ($f=0.6-0.7$), θ^* is in the range 28–30°, but for weak faults ($f=0.1-0.2$) (Carpinteri and Paggi, 2004) the range is 39–42°. It should be noted that faults can get reactivated in a wider range around the optimal angle. In the case of a caldera fault with normal strength and dipping inwards 80–85°, the opti-

mal σ_1 for reactivation would be in a plane perpendicular to the fault plane with dip 50–57°. For the same dipping of a weak fault, the dip range of σ_1 would be 38–46°.

Today most authors accept Anderson's derivation (1936) that cone sheets are mode I fractures formed by upward pressure of magma. He showed that they propagate in the direction of σ_1 and open up in direction of σ_3 . Sheets in the dip range 38–46° or 50–57° would therefore signify a paleo-stress field that was optimal to reactivate near vertical inward dipping (80–85°) weak or normal strength faults, respectively, according to the relation of Sibson (1985). Cone sheets are commonly observed within eroded central volcanoes in Iceland and are closely related to calderas. They have a wide range of dip. However, it seems common among all observations carried out, that peaks in distributions of cone sheet dips are within the range 25–45° (Annells, 1968; Sigurdsson, 1970; Jóhannesson, 1975; Franzson, 1978; Gudmundsson, 1998a; 2002; Siler and Karson, 2009; Burchardt *et al.*, 2011). Most authors also find steeper dipping sheets within the range ~60–90° (Annells, 1968; Jóhannesson, 1975; Franzson, 1978; Gudmundsson, 1998a, 2002; Siler and Karson, 2009; Burchardt *et al.*, 2011). The steep dipping sheets tend to be less numerous than the shallow ones, with one or two exceptions (Gudmundsson, 1998a; Franzson, 1978). Gudmundsson (1998b) has modelled the formation of normal fault calderas numerically. He predicts that during the time of doming of a magma chamber, σ_1 has intermediate dips (~30–45°) in the vicinity of the lower (deeper) half of the caldera fault, but in the upper half steepening (~40–75°) is indicated.

Distribution of cone sheets in Iceland therefore indicates paleo-stress field within extinct central volcanoes that may have been commonly favourably oriented to reactivate weak steeply dipping (80–85°) caldera faults, during periods of cone sheets formations. The assumed large ratio of slip to fault length of mature calderas in the world, and observations of rapid subsidence of caldera floors (e.g. Hartley and Thordarson, 2012; Sigmundsson *et al.*, 2015), suggests that caldera faults are commonly weak faults. As cone sheets dips in the range ~50–60° are by

no means uncommon in Iceland, the requirement of weak faults may not be necessary in order to reactivate steeply dipping normal caldera faults during periods of cone sheet formations. However, when the least effective principal stress is tensile, reactivation of regular strength high angle normal faults becomes easier (Sibson, 1985). This should be the situation expected in plate spreading environment like Iceland.

Studies on calderas in Iceland indicate a major caldera ring fault, with relatively regular circular or oval geometry, but the caldera floor has often considerable faulting and flexing (H. Jóhannesson, *pers. comm.*, Jan. 2015). However, some of the central volcanoes have more complex structure, e.g. couple of calderas within their domain (Jóhannesson and Sæmundsson, 2009). Therefore, Icelandic calderas are usually neither a pure end member piston collapse nor a chaotic piecemeal collapse on random faults, but comprise probably components of both. It is thought that caldera formations in Iceland take thousands of years to develop (Fridleifsson, 1973; Jóhannesson, 1975; Torfason, 1979; Franzson, 1978; Fridleifsson, 1983). However, there are observations, which indicate that incremental caldera collapse can be rapid (Hartley and Thordarson, 2012; Sigmundsson, 2015), the final adjustment of an incremental collapse taking half a century (Hartley and Thordarson, 2012).

Dynamics of the 1973–1996 Bárðarbunga earthquake sequence

If the caldera fault dip towards the centre of Bárðarbunga, which is the only evidence available from the geological record as of today, then thrust earthquakes on the caldera fault were caused by uplift movement, and the driving force within Bárðarbunga was likely to be increased pressure within the volcano (Figure 5; Bjarnason and Þorbjarnardóttir, 1996). The other explanation, with outward dipping caldera fault, would be decreased pressure with subsidence (Einarsson, 1991). Reactivated normal faults have been observed in Iceland (Gudmundsson *et al.*, 2008), and in the present work it is proposed that the observation of Fridleifsson (1983) should be interpreted as a reactivated inward dipping ring fault. Nettles and Ekström (1998) assume downward movement on outward dipping cone (ring) fault structure below an expanding

shallow magma chamber. Tkalčić *et al.*, (2009) propose two models for the 1996 event: a) a complex magma chamber, where volume decreases at the bottom of the chamber, but increases at the top of it, or b) volume loss in a magma chamber by opening of a dyke above it. Both models of Tkalčić *et al.*, (2009) are constrained with no net volume change solution of the moment tensor under Bárðarbunga, with a 2/3 part of the moment as non-double compensated-linear-vector-dipole.

The renewed activity in Bárðarbunga in the second half of the year 2014 does give a hint of the driving force of the earthquake sequence in 1973–1996. In the 2014 episode, GPS measurements of Bárðarbunga volcano show high rate vertical subsidence of the caldera floor, ~ 1.0 m/day during the first few weeks of the episode (Sigmundsson *et al.*, 2015). All moment tensor solutions of a series of intermediate size earthquakes within the Bárðarbunga 2014 episode, calculated from data recorded on international seismic networks, show predominantly normal faulting, with a large non-double-couple component (Global Moment Tensor Program by Ekström *et al.*, 2012; GEOSCOPE by Vallée *et al.*, 2011; GEOFON Program, 2014). These observations suggest a correlation between the non-double-couple normal faulting and caldera floor subsidence. The waveform characteristics of the current intermediate earthquakes are the same as previously described for the 1973–1996 earthquake sequence, except presumably for the direction of motion as determined by the international moment tensor solutions (e.g. the seismic station BORG IRIS/IDA in West Iceland, of the Global Seismic Network). Therefore, it is concluded that similar fault patches are moving in the 2014 episode as in the 1973–1996 sequence, but with opposite sense of motion. This supports the hypothesis that the 1973–1996 sequence was due to uplift of the caldera block, possibly piecemeal uplift of different parts of the block, due to increased pressure inside the volcano.

Einarsson (1991) proposed magma deflation in Bárðarbunga to be the cause of the thrust earthquake sequence. He observed a correlation between the Bárðarbunga main events and magma activity during the 1975–1984 volcanic episodes of the Krafla

central volcano, located 110 km north of Bárðarbunga. He proposed a pressure connection between the two volcanoes, along a hypothesised partially molten layer under Iceland. Magma flow into the Krafla magma chamber would thus cause pressure decrease and eventual collapse of the caldera floor in Bárðarbunga. The volcanic inflation of Krafla ceased in 1984, but the Bárðarbunga events continued until 1996, undermining the proposed mechanism. An alternative deflation model can be suggested in which the magma reservoir of Bárðarbunga is filled with magma from the mantle and partially emptied with subsurface lateral magma ejection, occurring periodically for 22 years, until Sept. 1996, when it reached the surface through weak zones of the region. An argument against this hypothesis is lack of observation of clear intrusion tremors, or other seismic activity that can be associated with magma injection into neighbouring regions before or after each of the Bárðarbunga main events. Such major seismic activity was only observed after the 1996 main event.

The driving force of an inflation model is ascending magma from the mantle that gradually saturates the storage capacity of the magma reservoir under the volcano. However, instead of a dominant lateral magma ejection when critical pressure is reached inside the magma chamber, the pressure lifts the caldera block. The main earthquakes occur when cylindrical faults (e.g. the caldera fault), that dip to the centre of the volcano, fail (Bjarnason and Þorbjarnardóttir, 1996). Immediately following the earthquake the pressure is decreased due to the increased volume of the volcano. The magma does therefore probably not flow out of the volcano in large quantities unless a dyke intrusion opens up volume outside the volcano, to the surface or subsurface, or if there is a relatively quick pressure increase after the earthquake. Such an increase in pressure can result from gas bubbles, rising up through the magma, causing a sudden pressure increase in the magma chamber (Linde *et al.*, 1994). Increased pressure of this kind could explain the hypothesised flow of magma out of magma saturated Bárðarbunga volcano, following the main earthquake of 1996. Neither the inflation nor deflation models do, however, explain the contrast in after seis-

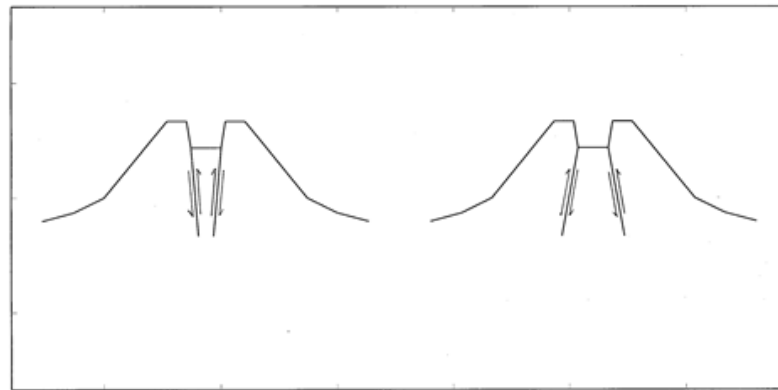


Figure 5. Two possible fault movements that are consistent with the observed focal mechanisms and moment tensors of the Bárðarbunga main events. To the left is an inflation model (resurgent caldera) with inward dipping caldera fault, and to the right is a deflation model with outward dipping caldera fault. Identification of the fault(s) that slip within the volcano in the main events is uncertain. Highly simplified tectonic picture of a central volcano with caldera is depicted, e.g. field observations in Iceland find caldera faults with near vertical dip. – *Mögulegar hreyfistefnur á aðalmisgengi Bárðarbunguöskjunnar, sem báðar geta skýrt brotlausnir og vægisþínur megin skjálfta í Bárðarbungu. Líklegast eiga megin skjálftar Bárðarbungu upptök á hringlaga öskjumisgengi. Tvær ólíkar túlkanir á orsökum skjálftanna koma til greina. Ef misgenginu hallar inn á við, er öskjuris líkleg orsök þeirra, en ef misgenginu hallar út, er líkleg orsök öskjusig. Athugið, að hér er dregin upp mjög einfölduð mynd af tectóník megineldstöðvar með öskju. Jarðfræðilegar athuganir á Íslandi sýna t.d. að halli öskjumisgengja er nærri lóðréttur.*

mic activity of the 1996 event in the caldera region and lack of such activity in previous main events. It can only be speculated that in 1996 the loading force had reached a maximum, and that the stress release was less complete in the 1996 main event compared to previous events, or that the ring fault had become significantly weakened compared to before, possibly due to magma lubrication with formation of a ring dyke or cone sheets entering the fault zone. In case of such complex geological events, the seismicity following the 1996 main event consists only partly of true aftershocks in the caldera region.

It is useful to compare source parameters of the Bárðarbunga events, recorded by the ICEMELT broadband seismic network in 1994–1996, with a couple of other events in the Iceland region recorded by the same network, as well as the larger 1987 Vatnafjöll event ($M_W=5.9$) (Bjarnason and Einarsson, 1991; Figure 1), recorded on a broadband seismograph installed by the Carnegie Institution of Washington, USA, in Akureyri North Iceland (Evans and Sacks, 1980).

One source parameter to be considered is the corner frequency of the seismic wave spectra, as the Bárðarbunga events have anomalously low frequency content or low corner frequency. In the Brune (1970) earthquake model, corner frequency is proportional to rupture velocity, which in turn is proportional to the S-wave velocity of the ruptured material (Scholz, 1990). Corner frequency is also frequently related to the concept dynamic stress drop of earthquakes, with low dynamic stress drop correlating with low corner frequency and slow rupture velocity. A vast literature exists on that subject of earthquake stress drop. Several authors have warned that there is a non-unique relation between static stress drop (equation 1 in Appendix) and dynamic stress drop, derived from the corner frequency of the earthquake spectra (Scholz, 1990; Atkinson and Beresnev, 1997). Stress drop estimate for an earthquake can differ by a factor of 4 to 5, especially if dynamic stress drop estimates are mixed with static stress drop estimates.

From the observations presented on long source duration and low corner frequency, it is concluded

that Bárðarbunga medium size earthquakes are highly anomalous earthquakes due to their low stress drop (see discussion in Appendix). There may be several explanations to this in the case of the Bárðarbunga intermediate size earthquakes. The shallow depth of the Bárðarbunga events is probably one of the principal factors for their low stress drop. It follows from equation 1 (in Appendix), that for two earthquakes with equal moment, the earthquake in higher shear modulus regions (i.e. usually relatively deeper source), would tend to have higher stress drop, given same fault geometry. At shallow depth the material friction is smaller than at greater depth, and hence the loading force or stress needed for slip is smaller, but loading force is proportional to stress drop (Scholz, 1990). Low stress drop events may have relatively longer principal fault dimension than a higher stress drop event of same magnitude, and or lower slip. The slip of the 1973–1996 sequence events is not constrained, but the ~ 12 km fault length of the 1996 event is comparable to the larger Vatnafjöll event (Figure 1). The reason for the difference in stress drop character between the Vatnafjöll and Bárðarbunga events is perhaps best explained by different loading mechanism. In the former case it is plate tectonic force acting on the entire crust, which probably holds strongest in the brittle part of the lower crust. In the latter, however, magma buoyancy acts upon the relatively weak upper crust, causing low stress drop earthquakes. This may also explain the small aftershock activity in most of the Bárðarbunga events; stress relaxation may be more complete within this low stress environment.

The shallow depth of the fault slip of the Bárðarbunga events suggests that the loading force is also shallow, like an increased pressure in shallow magma chamber postulated by Nettles and Ekström (1998). However, if the loading force originates from a greater depth, the Bárðarbunga events may signify breaking of shallow asperities on deeper extending well lubricated faults. This has been discussed by Das and Kostrov (1986). Therefore, it is not certain whether the loading force of the shallow Bárðarbunga intermediate size earthquakes originate from a shallow magma chamber in the upper and/or middle crust, or from a deeper magma reservoir in the lower crust.

CONCLUSION

The 1973–1996 Bárðarbunga sequence of intermediate size earthquakes is interpreted as being magmatic induced, caused by mantle derived magma seeping into the volcano. It led to increased pressure and lift of the caldera block with reactivated slip on shallow ring fault patches dipping towards its centre. On Sept. 29–30th 1996, the pressure inside the magma reservoir exceeded the lithostatic pressure, probably causing lateral dyke formation resulting in large magma pressure increase in the neighbouring region for the first time in the 1973–1996 earthquake sequence. The increased pressure led to volcanic eruption on the subglacial volcanic ridge Gjálpi, in NW Vatnajökull on Sept. 30th. The eruption may have caused a large pressure drop in Bárðarbunga and neighbouring regions, judged by lowered seismicity in the NW Vatnajökull area during the 8 years following the eruption.

The loading force of the shallow Bárðarbunga main events may be due to increased pressure in a shallow magma chamber, or it may be due to increased magma pressure at greater depth.

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ÁGRIP

Meðalstór jarðskjálfti ($M_W=5,6$) átti upptök sín undir eldfjallinu Bárðarbunga 29. september 1996. Þótt þessi jarðskjálfti teljist aðeins vera meðalstór miðað við jarðskjálfta yfirleitt í heiminum, þá var þetta stór skjálfti, þegar tekið er tillit til stærðar Bárðarbungu. Hann gæti hafa fært 12 km langa jarðspildu til um ~ 65 cm. Mikil skjálftavirkni fylgdi í kjölfar megin-skjálftans. Einum og hálfum degi eftir megin-skjálftann varð eldgos undir Vatnajökli, 20 km suðsuðaustur frá upptökum skjálftans, miðja vegu milli Bárðarbungu og Grímsvatna. Hefur eldgos þetta verið nefnt Gjálpargos. Vegna nálægðar þessara atburða í tíma og rúmi er freistandi að athuga, hvort orsakatengsl séu þarna á milli. Skjálftinn árið 1996 var síðasti skjálfti í röð meðalstórra og minni skjálfta, sem hófust í Bárðarbunga árið 1973. Þessi skjálftaröð hefur einkennst af nærri árlegum aðalskjálftum af stærðinni 4,5–5,7 m_b . Lág bylgjutíðni þeirra bendir til grunnra upptaka (≤ 5.0 km) og óvenju lítills spennufalls (<10 bör) miðað við tektoníska skjálfta. Lág spennufall skýrist af grunnnum upptökum í efri lögum jarðskorpunnar miðað við stærð skjálftanna. Þar er styrkur misgengja í brotgarðri skorpu minni en á meira dýpi. Brotlausnir og vægisþínur (e. moment tensor) sýna samgengishreyfingar á sveigðum misgengjum, annaðhvort vegna aukins eða minnkaðs þrýstings í eldfjallinu. Hér eru leidd rök að því, að skjálftinn hafi orðið vegna viðsnúinnar hreyfingar á öskjumisgengi, sem hallar inn (þ.e. samgengishreyfing á siggengi). Af því leiðir, að aukinn þrýstingur sé líklegasta orsök, og að þrýstingsaukningin stafi af flæði kviku inn undir Bárðarbungu. Í ljósi þessarar tilgátu hefur Bárðarbungufjallið verið að þenjast út í u.þ.b. aldarfjórðung, og í lok september 1996 náði þenslan hámarki, og kvika braut sér leið undan fjallinu. Eldgosið í Gjálpi hefur væntanlega valdið þrýstingslækkun í Bárðarbunga og nágrenni, sem sést í lítilli skjálftavirkni í norðvesturhluta Vatnajökuls frá miðju ári 1997 til 2005 (þ.e. engir skjálftar af stærðinni $>3,0$).

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APPENDIX

Stress-drop of Bárðarbunga earthquakes relative to regular tectonic earthquakes in Iceland.

Seismic stress drop is a factor that can account for difference in corner frequency of earthquakes of similar size. *A priori*, the corner frequency of e.g. the larger, Vatnafjöll event, is expected to be lower than that of the smaller Bárðarbunga 1996 event. It turns out that corner frequency of the Vatnafjöll event is three times as high as the Bárðarbunga event of 1996 (Table 2). Similarly the corner frequency of the larger Skagafjörður 1994 event ($M_W=5.5$) is twice as high as that of Bárðarbunga 1995 event ($M_W=5.4$) (Table 2). The Vatnafjöll earthquake nucleated in the lower crust, with centroid depth of 6.6 km (Bjarnason and Einarsson, 1991), but it is argued here that the unusually low corner frequencies of the Bárðarbunga medium size earthquakes do suggest shallow sources as previous authors have suggested (Einarsson, 1991; Nettles and Ekström, 1998; Konstantinou *et al.*, 2003; Tkalčić *et al.*, 2009).

Categorizing stress drop in Bárðarbunga earthquakes, three principal methods come to mind: Static stress drop estimates; dynamic stress drop from source time functions determined with moment tensor inversion; and dynamic stress drop from corner frequency:

Static stress drop

Calculation of static stress drop $\Delta\sigma_{static} = C_g\mu D/A^{1/2}$ (eq. 1) is not feasible because of lack of information on most of the parameters that it depends on, fault area (A), fault geometry (C_g), and average slip (D). However, the shear modulus (μ) in the fault region is assumed to be equal to the average modulus for the Central and North Iceland Volcanic zones (Bjarnason and Schmeling, 2009).

Dynamic stress drop and source time function

Nettles and Ekström (1998) reported unusually long source-time functions (4–7 s) for the intermediate Bárðarbunga earthquakes in the years 1976–1996, with 5 s for the 1996 event. Fichtner and Tkalčić (2010) concluded that the source duration of the 1996 event could not be well constrained, in spite of the higher frequency resolution of local broadband recordings used (HOTSPOT array, Foulger *et al.*, 2001). The duration in the range of 3–8 s was estimated by these authors, with maximum moment release in the first 3.5 s. Konstantinou *et al.* (2003) estimated ~ 5 s long source time function for the 1996 event. All these estimates suggest long source duration, and 5 s duration of the 1996 event

is $3/4$ longer than average for earthquakes of that size (Ekström *et al.*, 1992). Although longer than average source time function may be an indicator of low stress drop event, as the Brune model suggests, firm theoretical or empirical relations with observations are still lacking (see e.g. Scholz, 1990; Bizzarri, 2010). There is even a case of very long source duration event compared to the average that may not have been with low stress drop (Ekström *et al.*, 1992). It is, however, generally agreed that longer than average source duration indicates low rupture velocity (V_r).

For the estimated 12 km long rupture in the 1996 event, assuming unilateral rupture (reasonable assumption based on Stefánsson *et al.* (1996) aftershock distribution), the rupture velocity (maximum velocity) is 2.4 km/s, which is a low value. Assuming a normal value ratio of rupture velocity to source shear velocity to be $V_r/\beta = 0.9$, this gives source depth of 2.0–2.5 km, and $V_r/\beta = 0.7$, a source depth of ~ 5.0 km, using the shear velocity structure of Bjarnason and Schmeling (2009) for Central Iceland. It is not reasonable to assume V_r/β to be lower than 0.7, because that would place the source at unreasonable depth in the lower crust or even in the mantle. As stress drop (eq. 1) is a linear function of the shear modulus there is an indication that a shallow source earthquake would tend to have lower stress drop. In the laboratory this effect is observed: At low confining pressure (equivalent to shallow depth) the material friction is smaller than at greater pressure, and hence the loading force (stress) needed for slip is smaller. As stress drop is proportional to the loading force (Scholz, 1990), it follows that the stress drop is also lowered. Therefore, it is concluded, that these observations do indicate a low stress drop of a shallow (2.0–5.0 km) event. It seems unlikely that the rupture depths of the other Bárðarbunga events, with similar source properties, would deviate much from the above depth range.

Dynamic stress drop and corner frequency

The relationships of corner frequency f of earthquake spectra of a circular crack model of Sato and Hirasawa (1973) were reviewed by Aki and Richards (1980) [p. 820–821]. The Sato and Hirasawa (1973) model spectra have a Brune (1970) like ω^{-2} asymptote beyond the corner frequency. The dependence of azimuthally averaged P-wave corner frequency $\langle f_p \rangle$, is $2\pi\langle f_p \rangle = C_p\alpha/R$ (eq. 2), where α is P-wave velocity at the source, R is radius of the circular crack, and C_p is a scaling constant that depends on the ra-

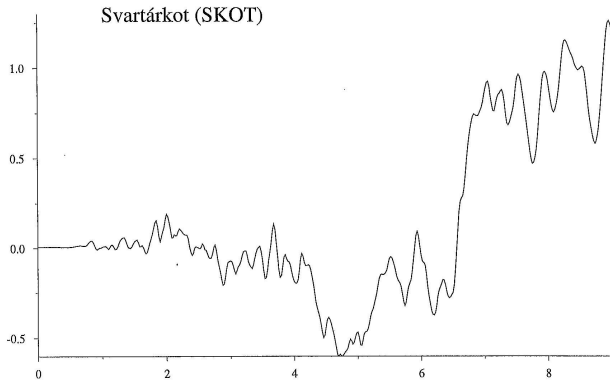


Figure A1. Waveform of the Bárðarbunga main event on Sept. 29th, 1996, recorded on vertical component at ICEMELT broadband seismic station SKOT at epicenter distances 77 km. A 9 second long window is shown. – Bylgjugögn meginskjálfta í Bárðarbungu 29. sept. 1996, skráð á lóðréttan þátt stöðvar í Svartárkoti í ICEMELT breiðbandsskjálftanetinu í 77 km fjarlægð frá skjálftaupptökum. Tímagluggi er 9 sekúndur. Sjá staðsetningar skjálftamæla á 1. mynd.

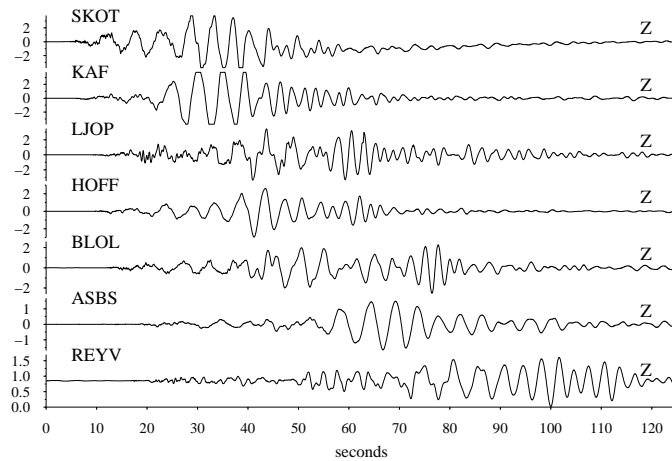


Figure A2. Waveforms of Bárðarbunga main event on Sept. 29th 1996, recorded at ICEMELT broadband seismic stations (see Figure 1 for station locations). Two minutes long, vertical component (Z) records are shown. Epicentral distances increase from the top record and down. – Bylgjugögn meginskjálfta í Bárðarbungu 29. sept. 1996, skráð á lóðréttan þátt stöðva í breiðbandsskjálftanetinu ICEMELT. Athugið, að skali á útslagi (y-ás) er breytilegur milli stöðva, og fjarlægð þeirra frá upptökum eykst frá efsta línuriti og niður.

tio Vr/β . Equivalent form exists for S-wave, but with different scaling constant. Combining the relation of moment and stress drop for a circular crack (e.g. Madariaga, 1980), $M_0 = \frac{16}{7} \Delta\sigma_{dyn} R^3$ (eq. 3), with eq. 2. gives $\Delta\sigma_{dyn} = \frac{7}{16} (2\pi/C_p\alpha)^3 M_0 \langle f_p \rangle^3$ (eq. 4). It is informative to evaluate what effect uncertainty in the centroid depth (or the Vr/β ratio) of Bárðarbunga events, or what affect the scale factor $C_p\alpha$ in eq. 4, might have on stress drop calculations in the Bárðarbunga region. The evaluation shows that variations in $C_p\alpha$ for plausible values of Vr/β in the range 0.7–0.9 do not (<2%) affect the stress drop significantly, therefore Vr/β is constrained at 0.9.

The dynamic stress drop (eq. 4) is calculated for a pair of tectonic earthquakes in Iceland and a pair of Bárðarbunga earthquakes, for which the moment magnitude is available (Table 2). The velocity models used for these calculations

are from Bjarnason and Schmeling (2009) for the Bárðarbunga events, and for the other events the standard earthquake location model for Iceland (SIL model), that was constructed from the work of Bjarnason *et al.* (1993). The stress drops of both Bárðarbunga earthquakes analysed here are exceptionally low, 5 ± 3 bars in 1994 and 4 ± 2 bars in 1996. At the time of the Vatnafjöll earthquake, there was only one broadband seismic station operating in Iceland; hence the corner frequency is estimated only from one direction, which increases the uncertainty of the estimate. However, based on the previous work of Bjarnason and Einarsson (1991) and calculation of 120 bars ($-60/+90$ bars) stress it is concluded that the Vatnafjöll earthquake was a high stress drop event. The event off-shore Skagafjörður in 1994 had 15 ± 2 bars stress drop, which is below global average stress drop but not unusual (Allmann *et al.*, 2009).